CrossMark
click for updates

RESOURCE LETTER

Resource Letter GECC-1: The Greenhouse Effect and Climate Change: Earth's Natural Greenhouse Effect

Stephen E. Schwartz

Environmental and Climate Sciences, Brookhaven National Laboratory, Upton, New York 11973

Earth's greenhouse effect is manifested as the difference between thermal infrared radiation emitted at the Earth surface and that emitted to space at the top of the atmosphere. This difference, which is due mainly to absorption and downward emission of radiant energy by atmospheric trace gases, results in global mean surface temperature about 32 K greater than what it would otherwise be for the same planetary absorption of solar radiation, and is thus of enormous importance to Earth's climate. This Resource Letter introduces the physics of the greenhouse effect and more broadly of Earth's climate system and provides resources for further study. A companion Resource Letter (GECC-2), planned for the following issue, examines the increase in the greenhouse effect due to human activities over the past 200 years and its consequences for Earth's climate system. © *2018 American Association of Physics Teachers.*
https://doi.org/10.1119/1.5045574

I. INTRODUCTION

The so-called "greenhouse effect" is essential to maintaining a climate that is hospitable to life on Earth as we know it. This Resource Letter describes Earth's natural greenhouse effect. A companion Resource Letter (GECC-2) examines the increase in the greenhouse effect due to human activity, denoted here as the *intensified greenhouse effect*. This intensified greenhouse effect, the major cause of the change in Earth's climate that has occurred over the industrial era, is expected to continue over at least the next several centuries. As the greenhouse effect exerts a major effect on Earth's radiation budget, it is necessary to quantify the several fluxes that contribute to this budget to provide context to the examination of changes in that budget due to the intensified greenhouse effect.

Before examining Earth's energy budget and the greenhouse effect, it may be instructive to consider a paradox regarding the energy budget of a human being. The nominal standard daily energy intake for an adult, 2000 Calories (kcal), corresponds to 8.4×10^6 J·day^{-1} or 100 W. Now consider a human being as a blackbody radiator; an emissivity of unity is a good approximation in the thermal infrared region of the electromagnetic spectrum. For a body temperature of $37\,°C$ or 310 K, the emitted radiative flux density is about 500 W·m^{-2}, as given by the Stefan-Boltzmann radiation law, σT^4, where σ is the Stefan-Boltzmann radiation constant, $\sim 5.67 \times 10^{-8}$ W·m^{-2}·K^{-4}. Approximating the

human body as a cylinder and calculating the surface area of an adult as 1 m^2 (2-m high $\times$ 0.5 m circumference) results in an emitted power of 500 W, well more than the caloric intake. How can this be? While we are radiating 500 W, our warm environment is radiating back at us. Hence, the net rate of loss of energy by radiation is a small fraction of the gross emission rate calculated by the Stefan-Boltzmann law.

The phenomenon just described is analogous to what takes place in Earth's climate system. Shortwave solar radiation from the Sun transmitted through the largely transparent atmosphere reaches the Earth's surface where it is absorbed, heating the surface. As detailed below, the gross radiative flux density emitted by the surface of the planet (land, water) greatly exceeds, on global and annual average, the input radiative flux density received from the Sun, just as the thermal infrared radiation emitted by a person greatly exceeds his or her caloric intake. The radiation emitted by the Earth's surface in the thermal infrared, where the atmosphere is much less transparent, is absorbed by the atmosphere, which is thereby heated and in turn radiates, in part downward toward the surface. This *downwelling* infrared radiation from the atmosphere to the surface greatly diminishes the net upward infrared flux from the surface relative to the gross blackbody flux, diminishing the radiative cooling of the surface, thereby increasing the surface temperature, and resulting in the emitted thermal infrared radiation by the surface greatly exceeding the incoming solar radiative flux. This phenomenon is called the greenhouse effect, based on the erroneous

supposition that a greenhouse maintains its warmth by absorption and downward emission of longwave radiation from the glass surfaces.[1]

This Resource Letter examines the greenhouse effect of Earth's climate system; the companion Resource Letter (GECC-2) examines the intensified greenhouse effect, the increase in the greenhouse effect during the Anthropocene epoch. The term "Anthropocene," introduced by 1995 Nobelist for atmospheric chemistry Paul Crutzen,[2,3] is now widely used to characterize the epoch of Earth's geological history that began with industrialization some 200 years ago. Examination of the greenhouse effect requires considerable understanding of Earth's climate and the processes that control it. The focus here is physical processes. Although these physical processes cannot be wholly separated from chemical and biological processes, much insight is gained by looking at the climate system as a physical system. Much attention is directed to the global scale radiation budget and the greenhouse effect and its influence on surface temperature. GECC-2 examines perturbations in the global radiation budget due to the incremental amounts of atmospheric infrared-active gases (greenhouse gases or GHGs) resulting from human activities and resulting perturbations in global temperature.

Sources and References. A broad and detailed examination of present understanding of the climate system and climate change is given in the several Assessment Reports of Working Group I of the Intergovernmental Panel on Climate Change (IPCC). The most recent report, the Fifth,[4] was published in 2013; the Fourth report[5] occasioned award of the Nobel Peace Prize to the IPCC. These reports, which are freely downloadable on the web, are essential references for any student of climate and climate change. Several texts that might serve as an introduction to the climate and to the physics of climate are noted in Refs. 6–13. A hypertext history of the development of understanding of the greenhouse effect and climate change more generally by Weart,[14] which supplements his very readable book, may be accessed at http://www.aip.org/history/climate/index.htm. Citations to the primary and secondary literature given in this Resource Letter are intended to be of historical and pedagogical value, as well as to lead the reader to the research literature without attempting to be exhaustive. An earlier pedagogical reference[15] is also noted.

Geophysical data. Many data sets pertinent to climate and climate change are available on the internet, far too many to list exhaustively. NASA's Clouds and the Earth's Radiant Energy System (CERES) project[16] provides measurements of short- and longwave radiant fluxes at a variety of spatial and temporal scales. Measurement-based temporal and spatial data of Earth's surface temperature are maintained by NASA, NOAA, and the Climatic Research Unit of the University of East Anglia.[17–19] Meteorological reanalysis data are available from NOAA[20] and the European Center for Medium Range Weather Forecasts.[21] The World Data Center (WDC) for Paleoclimatology[22] maintains archives of ice core data from polar and low-latitude mountain glaciers and ice caps throughout the world. Additional data sets pertinent to climate change over the Anthropocene epoch are noted in Resource Letter GECC-2.

Organization of this Resource Letter. Section II presents an overview of Earth's radiation budget and emphasizes that to good approximation Earth's climate system can be characterized as a steady state. Section III provides a brief overview of Earth's climate system with emphasis on thermal structure of the atmosphere as a function of altitude pertinent to the greenhouse effect. Section IV deals with the role of the greenhouse effect on Earth's climate and identifies the key GHGs.

Resource Letter GECC-2 introduces the concept of radiative forcing of climate change, examines increases in GHGs over the Anthropocene epoch and the resultant radiative forcings, describes climate system response to these forcings, introduces the climate sensitivity concept that relates forcings to observed temperature changes, and examines implications for prospective future climate change.

Supplementary Notes (SN) in the online version of this Resource Letter provide detail or lend perspective, as follows.

SN1. The consequences of a global energy imbalance
SN2. Why are water vapor and carbon dioxide strongly infrared-active whereas nitrogen and oxygen are not? The roles of quantum mechanics
SN3. Correlations between climate properties and greenhouse gases over the glacial ice ages

1. "Note on the Theory of the Greenhouse," R. W. Wood, Philos. Mag. **17**, 319–320 (1909). https://www.tandfonline.com/doi/abs/10.1080/14786440208636602?journalCode=tphm17. Shows that actual greenhouses maintain their warm temperature mainly by the enclosure's suppression of convective transport of warm air away from the surface. Historical and popular interest; might serve as basis for laboratory experiment. (E)

2. "The Anthropocene," P. J. Crutzen and E. F. Stoermer, IGPP Newsletter No. **41**, pp. 16–18 (2000). http://www.igbp.net/download/18.316f18321323470177580001401/1376383088452/NL41.pdf. Historical. (E)

3. "Geology of mankind," P. J. Crutzen, Nature **415**, 23–23 (2002). https://www.nature.com/articles/415023a. Historical and influential commentary. (E)

4. **Climate Change 2013**: **The Physical Science Basis (AR5)**, T. F. Stocker *et al.*, Editors (Cambridge U.P., Cambridge, United Kingdom and New York, NY, 2013). http://www.climate2013.org/images/report/WG1AR5_ALL_FINAL.pdf. Intergovernmental Panel on Climate Change. Essential reference. (E)–(I)

5. **Climate Change 2007**: **The Physical Science Basis (AR4)**, S. Solomon *et al.*, Editors (Cambridge U.P., Cambridge, United Kingdom and New York, NY, 2007). http://www.ipcc.ch/publications_and_data/ar4/wg1/en/contents.html. Intergovernmental Panel on Climate Change. Essential reference. (E)–(I)

6. **Global Warming**: **The Hard Science**, L. D. D. Harvey (Prentice Hall, London, 2000). 336 pp., ISBN 0582-38167-3. Very approachable introduction to climate change. (E)–(I)

7. **Global Warming: The Complete Briefing**, 5th edition Houghton J. (Cambridge U.P., Cambridge, 2015) 380 pp. ISBN: 9781107463790. Thorough overview of climate and climate change; strong on principles; with figures and tables and no equations. (E)–(I)

8. **Dynamic Climatology: Basis in Mathematics and Physics**, J. N. Rayner, (Blackwell, Malden MA, 2001) 304 pages ISBN: 978-1-577-18016-6. Introductory with a review of underlying mathematics, statistics, geometry, thermodynamics, mechanics, radiation. (E)–(I)

9. **The Physics of Atmospheres, Third Edition**, J. T. Houghton (Cambridge University Press, Cambridge, 2002). 320 pages. ISBN 0-521-01122-1. Good starting point for physics students wishing to delve into atmospheric physics. (E)–(A)

10. **Physics of Radiation and Climate**, M. A. Box, and G. P. Box, (CRC Press, 2015) 495 Pages ISBN 9781466572058. Introduction to the physics of radiation and radiation transfer, with emphasis on the central role of radiation in Earth's climate. Mathematical. (I)

11. **Principles of Planetary Climate**, Raymond T. Pierrehumbert (Cambridge, 2010). 674 pages ISBN: 9780521865562. Strong on physical principles. Mathematical. (I)–(A)

12. **An Introduction to Atmospheric Radiation, Second Edition**, K. N. Liou (Academic Press, 2002) 583 Pages ISBN: 9780124514515. Treats not only atmospheric radiation but also the properties of Earth's atmosphere that influence radiation transfer; also contains an extensive treatment of climate change. (E)–(A)

13. **Physics of Climate**, J. P. Peixoto and A. H. Oort (Am. Inst. of Phys., New York, 1992) 520 pages ISBN 978-0-88318-712-8. Another good starting point for physics students wishing to delve into atmospheric physics. (E)–(A)

14. **The Discovery of Global Warming: Revised and Expanded Edition**, Spencer R. Weart (Harvard U.P., 2008) 240 pages ISBN 9780674031890. Good introduction to climate, greenhouse effect, and intensified greenhouse effect. (E)

15. "An introduction to global warming," John R. Barker and Marc H. Ross, Am. J. Phys. **67**, 1216 (1999). Covers many points treated here, but in less detail; good introduction. (E)

16. The Clouds and the Earth's Radiant Energy System (CERES). https://eosweb.larc.nasa.gov/project/ceres/ceres_table

17. GISS Surface Temperature Analysis (GISTEMP). https://data.giss.nasa.gov/gistemp.

18. NOAA Merged Land Ocean Global Surface Temperature Analysis. https://www.ncdc.noaa.gov/data-access/marineocean-data/noaa-global-surface-temperature-noaaglobaltemp

19. HadCRUT4 global temperature dataset. https://crudata.uea.ac.uk/cru/data/temperature

20. Meteorological reanalysis data from NOAA National Centers for Environmental Predictions. https://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis2.html

21. Data sets from European Centre for Medium Range Weather Forecasts including reanalysis products. http://apps.ecmwf.int/datasets/

22. The World Data Center (WDC) for Paleoclimatology, https://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/ice-core

II. OVERVIEW OF EARTH'S RADIATION BUDGET

A. Top-of-atmosphere budget

Earth's climate, a dynamic, chaotic, near-steady-state system, is driven almost entirely by absorption of solar energy. This energy heats the planet and is responsible for climate and weather: wind, ocean currents, rain and snow, and temperature. "Near steady state" means that the energy introduced into the system is balanced, almost exactly, by energy leaving the system. Here "almost exactly" means that the difference is very small relative to the quantity of interest. Satellite measurements confirm, on global and annual average, the near equality of the absorbed power (from the Sun), denoted as *shortwave energy*, and the emitted power (in the thermal infrared), denoted as *longwave energy*. Conveniently, these two spectral ranges, calculated according to the *Planck radiation law* for black bodies [Ref. 23; Max Planck, Nobel Prize 1918] at 5770 K and 255 K, respectively, are almost completely non-overlapping (Fig. 1). This lack of overlap simplifies the discussion and facilitates measurements of the different components. For the rate of absorption of solar energy normalized to the surface area of the planet, Q, and for the rate of emission of thermal infrared energy, similarly normalized, $-E$ (the sign convention is that positive denotes heat *into* the climate system), then to good approximation on global and annual average,

$$Q + E = 0. \tag{1}$$

That is, the system is approximately in steady state, with the magnitudes of the individual terms greatly exceeding the net rate of energy uptake or loss.

The rate of energy absorption, Q, is very nearly equal to one fourth of the so-called "solar constant," J_S, times the planetary co-albedo, γ,

$$Q = \gamma J_S/4. \tag{2}$$

The factor of 1/4 is geometric, the ratio of the area of the planet to that of the subtended disk being almost exactly 4. The *co-albedo*, γ, is the complement of the planetary albedo, where the planetary *albedo* (more formally, Bond albedo[25]), is the fraction of incident solar irradiance that is reflected (scattered) back to space. Viewed from space, Earth is not black: clouds and deserts are bright; oceans and forests are dark, yielding a mean planetary albedo, determined by satellite measurements,[26,27] of about 0.29. The solar constant, J_S, is the flux density (hereinafter, flux) of solar irradiance at the mean Earth-Sun distance. For the measured value of J_S, about 1360 W·m^{-2},[28] the absorbed power Q is about 240 W·m^{-2}; the emitted longwave power is also about 240 W·m^{-2}.[27] Considering the planet as a black body in the thermal infrared, permits a radiative temperature, T_{toa}, at the
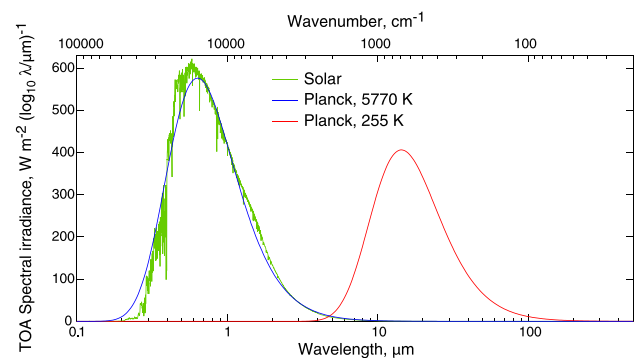
Fig. 1. Spectral distribution of downwelling shortwave and upwelling longwave irradiance at the top of the atmosphere using the Planck radiation law for black bodies at 5770 K (normalized to 340 W·m^{-2}) and 255 K versus wavelength or wavenumber (upper and lower axes, respectively, logarithmic scale). Also shown for comparison is measured spectral solar irradiance (Ref. 24) at mean Earth-Sun distance, divided by 4 to correspond to global-mean irradiance.

top of the atmosphere (TOA), to be calculated by the Stefan Boltzmann radiation law

$$E = -\sigma T_{\text{toa}}^4; \quad T_{\text{toa}} = (-E/\sigma)^{1/4}. \tag{3}$$

Solving for T_{toa} gives about 255 K, which is the temperature one would ascribe to the planet if measuring its thermal emission from space. Such a temperature would imply that Earth is a frozen planet.

The difference (32 K) between the radiative temperature of the planet, 255 K, and the global mean surface temperature (GMST, T_s), 287 K, is a consequence, or measure, of Earth's greenhouse effect. That the emitted surface flux exceeds the absorbed radiant energy was recognized by Fourier as early as the 1820s. Fourier correctly attributed the elevated temperature at Earth's surface in part to the lack of transparency of the atmosphere to thermal radiation, in contrast to its transparency to visible radiation.[29] This hypothesis was confirmed by Tyndall,[30] who established that downward emission of infrared radiation is due to trace gases in the atmosphere, principally water vapor and carbon dioxide. Tyndall drew the analogy that "As a dam built across a river causes a local deepening of the stream, so our atmosphere, thrown as a dam across the terrestrial rays, produces a local heightening of the temperature at the earth's surface." Thermal infrared radiation emitted at the surface is absorbed by infrared-active gases in the atmosphere and by clouds. These gases and clouds emit in the thermal infrared and some of the emitted power is returned to the surface of the planet, thus increasing the surface temperature to a value greater than it would be absent this greenhouse effect.

B. Global energy balance

As noted above, the global, annual average of the longwave irradiance emitted at the TOA is virtually equal to that of the absorbed shortwave irradiance. Why should this be? If for some reason the absorbed shortwave irradiance were to increase, planetary temperature would increase, increasing the outgoing longwave irradiance and restoring the balance. There is thought to be a very slight imbalance in the net irradiance at the TOA in natural climate system, even at steady state, due to non-radiative energy sources, mainly natural radioactive decay and the release of gravitational potential energy of the solid Earth; these and other, very minor terms in Earth's radiation budget are examined in Ref. 31. A contribution to energy imbalance of the planet has been imposed by changes in atmospheric composition over the past 250 years (examined in Resource Letter GECC-2). In principle, the net energy uptake by the planet, the difference between absorbed shortwave and emitted longwave radiation integrated over the planet

$$N \equiv Q + E = (J_S/4 - J_{\text{sw}\uparrow}) - J_{\text{lw}\uparrow} \tag{4}$$

might be measured from space. However, the energy imbalance is well less than the present accuracy of satellite measurements (about $\pm$ 3 W·m^{-2}),[27] so the assertion of global near equality of absorbed shortwave irradiance and emitted longwave irradiance rests more on the consequences of an imbalance, a heating or cooling of the planet, than on measurement (SN1). From measurements mainly of the change of ocean temperature over the past several decades, it is estimated that the net planetary energy imbalance is about

0.5 W·m^{-2} (Ref. 4, Box 3.1). Naturally caused fluctuations result from identified causes (variations in solar irradiance and short-lived increases stratospheric aerosols due to volcanic eruptions) and internal variability of the climate system (e.g., El Niño).

C. Earth's energy budget

Current estimates of global and annual means of the several fluxes that comprise Earth's energy budget, averages of quantities that vary greatly in space and time, are shown in Fig. 2. The global mean thermal infrared radiative flux emitted at the surface, calculated by the Stefan-Boltzmann law for a GMST 287 K with emissivity taken as unity, is 385 W·m^{-2}. One measure of the greenhouse effect is the difference between longwave flux emitted at the surface (385 W·m^{-2}) and that exiting at the top of the atmosphere, $\sim$145 W·m^{-2}. The radiative flux from the surface is augmented by a further 110 W·m^{-2} by transfer of latent heat (evaporation of water that condenses in the atmosphere and is returned to the surface as precipitation) and sensible heat (enthalpy, transferred by conduction and/or convection). Much of the power emitted at the surface is returned to the surface by downwelling longwave radiation from the atmosphere to the surface. Referring the emitted longwave energy flux at the TOA, $-E$, to the GMST, T_s, permits definition of an *effective emissivity* of the planet ε such that

$$-E = \varepsilon \sigma T_s^4. \tag{5}$$

For the surface emitted flux, 385 W·m^{-2}, and $-E = 240$ W·m^{-2}, the effective planetary emissivity, $\varepsilon = 0.62$. This emissivity is likewise a measure of the magnitude of Earth's greenhouse effect. These considerations establish the importance of the greenhouse effect in Earth's climate system.

Also shown in Fig. 2 are estimates of the one-sigma uncertainties in the several flux terms, although historically such uncertainty estimates have proved to be optimistic.[32] Based on these considerations, the climate system must be viewed
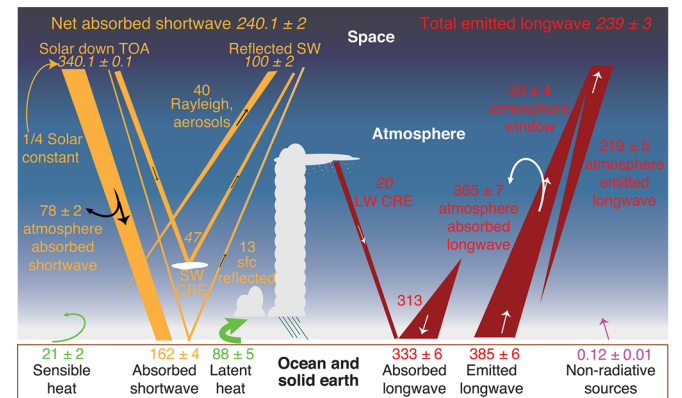

Fig. 2. Energy flows in Earth's climate system. Quantities shown are global and annual mean energy fluxes (W·m^{-2}) between the atmosphere and space, and between the atmosphere and the Earth surface. Satellite-measured quantities and associated uncertainties are in italics; remaining fluxes are calculated with 1-σ uncertainty estimates. Shortwave (solar) fluxes are shown in light gray (tan; color online); longwave (thermal infrared) fluxes in dark gray (red; color online); sensible and latent heat from surface to atmosphere in medium gray (green; color online); non-radiative sources (natural radioactive decay, gravitational collapse, fossil fuel combustion, nuclear energy) in medium gray (magenta; color online). TOA, top of atmosphere; SW, shortwave; LW, Longwave; CRE, cloud radiative effect; sfc, surface. Modified from Ref. 32.

as in very close energy balance at the top of the atmosphere, in the atmosphere itself, and in the upper, mixed layer of the world ocean.

Although depictions of global and annual mean energy fluxes such as Fig. 2 are useful to set the scene by showing the magnitudes of these fluxes, they do not display the large variability characteristic of these quantities. The several terms in Earth's energy budget exhibit substantial seasonal variation, up to $15 \, \text{W} \cdot \text{m}^{-2}$, even at global mean, a consequence mainly of the eccentricity of Earth's orbit, the obliquity of its rotational axis, and hemispheric asymmetry in the oceans. In contrast, the interannual variability is quite small, less than $1 \, \text{W} \cdot \text{m}^{-2}$ over a decade of high quality measurements from space.[27] Any changes in Earth's climate in response to externally imposed perturbations would be manifested in changes in the several radiation terms in addition to changes in surface temperature and ocean heat content. These seasonal variations, which are large compared to the forcings imposed on the climate system over the Anthropocene epoch, serve as further backdrop for examination of climate change and are indicative of measurement and modeling challenges that face the climate change research community.

Radiative fluxes other than at the TOA exhibit substantially greater uncertainties than fluxes at the TOA. The Stefan-Boltzmann law is used to calculate the longwave emission from the surface to the atmosphere, often using the global mean near-surface air temperature. Concerns over this approach are use of a global mean temperature versus more explicit consideration of spatial and temporal variation, use of the temperature of air (typically measured at $2 \, \text{m}$) rather than that of the actual radiating element, the surface, and the usually tacit assumption that the surface emissivity is unity.[33] Atmospheric absorption of longwave radiation is estimated by radiative transfer calculations that take into account not only absorption by the long-lived GHGs, whose mixing ratios are more or less uniform as a function of location in the atmosphere, but also absorption by water vapor, whose mixing ratio is highly variable spatially and temporally (Sec. IV B), and by clouds, also highly variable. Calculating the absorption and emission of longwave radiation by water vapor and clouds relies on representation of water vapor and cloud amount, vertical distribution, and temperature obtained from meteorological reanalyses; these reanalyses consist of weather forecast models run retrospectively, constrained by observations, using much the same physics as in climate models.[34–36] Surface absorption of shortwave irradiance is estimated based on surface spectral reflectance, which is dependent on surface type, soil moisture, vegetation cover, and the like, and on the downwelling spectral irradiance incident on the surface, which is greatly influenced by the presence and nature of clouds, and thus highly variable, spatially and temporally.

Further terms in the radiation budget are transfer of latent and sensible heat from the surface to the atmosphere. Latent heat release is estimated from the precipitation rate, as the global rate of net evaporation (evaporation minus dewfall) is equal to the global rate of precipitation. The latent heat flux indicated in Fig. 2 ($88 \pm 5 \, \text{W} \cdot \text{m}^{-2}$) corresponds to an annual and global mean precipitation rate of $1.1 \pm 0.06 \, \text{m} \cdot \text{yr}^{-1}$. Despite satellite-based observations, there is still substantial uncertainty in global precipitation. Recent work suggests that current estimates may be low by 10% or more.[37] Sensible heat flux is based on model calculations that take into consideration surface wind stress (which depends on surface roughness and wind speed) and temperature gradient between atmosphere and surface, all of which quantities can be highly variable spatially and temporally.

D. Spatial and temporal variability

At high time- and space-resolution, upwelling radiative fluxes at the TOA are much more variable than in global and longer-term means, Fig. 3. The figure illustrates the richness of the processes that govern absorption and emission of radiant energy from and to space. The local spatial structure is due largely to clouds, which are cold in the thermal infrared (because of the decrease in temperature with altitude) and bright in the shortwave. Notable is the intertropical convergence zone near the equator that is characterized by strong rising motion of the atmosphere is readily apparent in the Pacific. Contrast of bright land areas with adjacent ocean areas may be discerned. Likewise, the contrast between hotter land surfaces and adjacent oceans is readily discerned in the thermal infrared flux at the Persian Gulf. The figure illustrates the high dynamic range of the individual upwelling radiation terms, more than $1000 \, \text{W} \cdot \text{m}^{-2}$ in the shortwave and more than $300 \, \text{W} \cdot \text{m}^{-2}$ in the longwave. The instantaneous net daytime flux exhibits even greater dynamic range, $\sim 1300 \, \text{W} \cdot \text{m}^{-2}$. As this net flux is confidently thought to be less than $1 \, \text{W} \cdot \text{m}^{-2}$ on global, annual average, the large dynamic ranges of the several fluxes place stringent requirements on measurement accuracy. Although the instruments aboard the Sun-synchronous satellites that are the principal sources for Earth radiation budget data are quite accurately calibrated, current satellites sample only a limited portion of the diurnal cycle. Consequently, accounting for the full diurnal cycle requires that the rest of the diurnal cycle be filled in by measurements made with less-well calibrated instruments aboard geostationary satellites. Similar accuracy is thus required from those measurements and from models used to calculate diurnal averages. The results of such calculations are illustrated in panel *d*, which shows the smoothing that results from the summation of the several terms that is manifested in the reduced dynamic range of the data. This panel also illustrates the consequence of latitudinal transport of heat from equatorial regions (net absorption) to high latitudes (net emission).

A key strength of spatially resolved measurements is the ability to identify cloud-free regions and thus determine separately the irradiance from planet as a whole and from the cloud-free regions. The contribution of clouds to the short- and longwave irradiance components of the total upwelling flux, denoted cloud radiative effect, CRE, Fig. 2, is determined by difference.

E. Spectral dependence

Yet another important dimension essential to describing Earth's energy budget is the spectral dependence of both the short- and longwave upwelling radiation. An examination of the latter, shown in Fig. 4, is very instructive. This figure, which was obtained by an early infrared spectrometer in space,[38] shows the rich spectral structure associated with the upwelling longwave radiation, specifically for a location in a moist vegetated region in the Niger valley. For this location, the radiance at the TOA is bounded by a temperature profile that corresponds to the Planck function at $320 \, \text{K}$. In the spectral regions 8–9 and 10–12 μm, where absorption by water
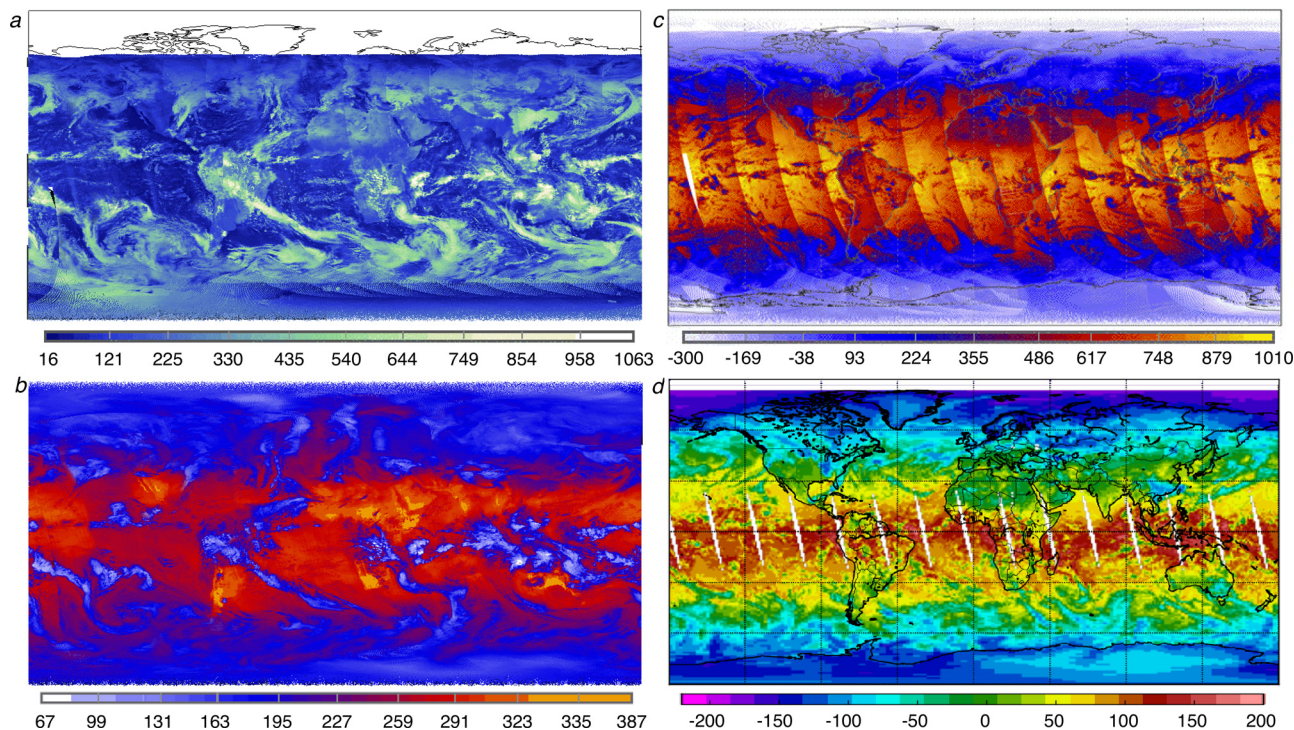
Fig. 3. Local top-of-atmosphere radiative fluxes (W·m^{-2}) as determined from satellite measurements, March 10, 2012. (a) Instantaneous shortwave reflected flux; (b) instantaneous longwave flux (positive upward); (c) instantaneous net daytime flux (positive downward), evaluated as $J_S \cos \theta_0$ minus the sum of upwelling SW and LW fluxes, where J_S is the solar constant and θ_0 is the solar zenith angle; (d) daily mean net flux (positive downward) after temporal integration. The depicted discontinuities are artifacts resulting mainly from changes, especially in cloudiness, between successive satellite overpasses. Data from the NASA CERES (Clouds and the Earth's Radiant Energy System) program; courtesy, Norman Loeb, NASA.

vapor, CO_2 and other trace atmospheric gases is slight, the radiance at the TOA is approximately equal to that given by the Planck distribution for the surface temperature; (as the measured quantity is radiance, its numerical value is less, by roughly a factor of π, than that of irradiance given in Fig. 1). These regions are denoted "window regions" of the thermal infrared. The radiation temperature in these window regions,
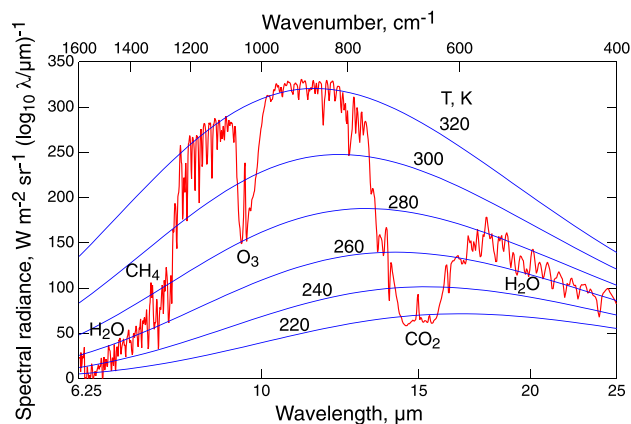

Fig. 4. Dependence of upwelling thermal infrared radiance on wavelength or wavenumber (lower and upper axes, respectively, logarithmic scale) measured by interferometry from the Nimbus 4 satellite platform. Spectrum, obtained on May 5, 1970, was from a moist, vegetated area of the Niger Valley (footprint roughly $100 \, km \times 100 \, km$, largely if not completely cloud free). Gases principally responsible for absorption are noted. Blue curves (color online) denote Planck distribution functions for indicated temperatures. Redrawn after Ref. 38; data from https://disc.gsfc.nasa.gov/datasets/IRISN4RAD_001/summary; acknowledgment to J. E. Johnson for assistance in acquiring the data.

320 K, is a fairly accurate measure of the local emission temperature of the surface, which substantially exceeds the local surface air temperature. Attention is called to the decrease in emitted radiation in the absorption bands of the GHGs water vapor, CO_2, ozone O_3, and methane CH_4, a manifestation of the greenhouse effect (Sec. IV). These features correspond to infrared-active transitions between vibrational states of the molecules (and, for water vapor, rotational states). The radiation temperature associated with the absorption bands of the several GHGs is a measure of the temperature of the atmosphere at which the gases exert their greatest influence on emission of radiation to space. The altitude z to which the temperature corresponds (by Fig. 5, Sec. III) is down from the TOA by approximately one unit of optical depth. The brightness temperature in Fig. 4 thus corresponds at each wavelength to the temperature at this altitude of peak-emitted radiance. The overall reduction in emitted radiance due to the presence of the GHGs is about 40%. That is to say, the effective TOA emissivity relative to the Stefan-Boltzmann radiance at the surface temperature is about 60%, consistent with the value obtained above.

In contrast to the strong absorption due to the infrared-active trace gases, N_2 and O_2, despite their much greater abundances, do not exhibit appreciable absorption in the thermal infrared, as electronic dipole vibrational transitions are strongly forbidden in homonuclear diatomic molecules because of quantum-mechanical symmetry considerations, **SN2**.

23. "Über das gesetz der energieverteilung im normal-spectrum," M. Planck, Ann. Phys. **309**, 553–563 (1900). English translation "On the law of distribution of energy in the normal spectrum," http://www.to.infn.it/

~zaninett/articoli/planck.html. Historically important. (I)

24. "The Sun's total and spectral irradiance for solar energy applications and solar radiation models," C. A. Gueymard, Sol. Energy **76**, 423–453 (2004). https://www.sciencedirect.com/science/article/pii/S0038092X03003967. Authoritative examination and tabular presentation of solar spectral irradiance. (E)–(I)

25. "Planetary radiation budgets," R. Kandel and M. Viollier, Space Sci. Rev. **120**, 1–26 (2005). Accessible introduction. (E)–(I)

26. "Seasonal variation of cloud radiative forcing derived from the earth radiation budget experiment," E. F. Harrison et al., J. Geophys. Res. **95**, 11679–11698 (1990). Research paper reporting measurements from the NASA Earth Radiation Budget Experiment. (E)–(I)

27. "Toward optimal closure of the Earth's top-of-atmosphere radiation budget," N. G. Loeb et al., Climate **22**, 748–766 (2009). Research paper reporting TOA fluxes determined by the NASA CERES program and associated error budget. (I)

28. "A new, lower value of total solar irradiance: Evidence and climate significance," G. Kopp and J. L. Lean, Geophys. Res. Lett. **38**, L01706 (2011). http://onlinelibrary.wiley.com/doi/10.1029/2010GL045777/full. Research paper. (I)

29. "Joseph Fourier, the 'greenhouse effect,' and the quest for a universal theory of terrestrial temperatures," J. R. Fleming, Endeavour **23**(2), 72–75 (1999). https://www.sciencedirect.com/science/article/pii/S0160932799012107 Historical and pedagogical review. (E)

30. "On the absorption and radiation of heat by gaseous matter, Second Memoir," J. Tyndall, Philos. Trans. R. Soc. London **152**, 59–98 (1862) http://rstl.royalsocietypublishing.org/content/152/59.short. Historical memoir. (I)

31. "Where does Earth's atmosphere get its energy?," A. C. Kren, P. Pilewskie, and O. J. Coddington, Space Weather Space Clim. **7**, A10 (2017). Systematic examination of sources of Earth's energy. (E)–(I).

32. "Observing and modeling Earth's energy flows," B. Stevens and S. E. Schwartz, Surv. Geophys. **33**, 779–816 (2012) https://link.springer.com/article/10.1007/s10712-012-9184-0. Review and assessment. (I)

33. "Earth's global energy budget," K. E. Trenberth, J. T. Fasullo, and J. Kiehl, Bull. Am. Meteorol. Soc. **90**, 311–323 (2009) http://journals.ametsoc.org/doi/abs/10.1175/2008BAMS2634.1. Accessible examination of observations, models, theory. (E)–(I)

34. "The NCEP/NCAR 40-year reanalysis project," E. Kalnay et al., Bull. Am. Meteorol. Soc. **77**, 437–471 (1996). http://journals.ametsoc.org/doi/abs/10.1175/1520-0477(1996)077%3C0437:TNYRP%3E2.0.CO;2. Historically important introduction to meteorological reanalysis. (E)–(I)

35. "The NCEP climate forecast system reanalysis," S. Saha et al., Bull. Am. Meteorol. Soc. **91**, 1015–1058 (2010). http://journals.ametsoc.org/doi/abs/10.1175/2010BAMS3001.1. Description of the NOAA National Centers for Environmental Predictions project and introduction to the data. (E)–(I)

36. "The ERA-Interim reanalysis: Configuration and performance of the data assimilation system," D. P. Dee et al., Quart. J. R. Meteorol. Soc. **137**, 553–597 (2011). https://onlinelibrary.wiley.com/doi/abs/10.1002/qj.828.

Introduction to Reanalysis by European Centre for Medium Range Weather Forecasts. (E)–(I)

37. "An update on Earth's energy balance in light of the latest global observations," G. L. Stephens et al., Nat. Geosci. **5**, 691–696 (2012). Review and assessment. (I)

38. "The Nimbus 4 infrared spectroscopy experiment 1. Calibrated thermal emission spectra," R. A. Hanel et al., J. Geophys. Res. **77**, 2629–2641 (1972). http://onlinelibrary.wiley.com/doi/10.1029/JC077i015p02629/full. Early infrared spectroscopic measurements from space. (E)–(I)

III. OVERVIEW OF EARTH'S CLIMATE

The physical laws governing Earth's climate are few, but powerful:

- *Conservation of matter*, reflected in the continuity equation;
- *Conservation of energy*, in all its forms: radiant, kinetic, potential, sensible, and, importantly, latent heat associated with phase transitions of water;
- *Newton's second law*, represented by the Navier-Stokes equation relating the change in momentum (of parcels of air or water) to (principally) gravitational and viscous forces;
- The *Planck radiation law* for black bodies;
- *Kirchhoff's law* equating spectral absorptivity and emissivity; and
- The *ideal gas law*.

To these physical laws must be added *material properties*, such as the fluidic, thermal, and spectral properties of water and *Earth-specific properties* such as the amount of water in the system, the salinity of the ocean, and the composition of the atmosphere. Important *boundary conditions* are the spectral irradiance of the Sun, orbital geometry, Earth's diameter and rotational speed, the shapes of continents, and the elevation of the land surface and the ocean floor. Also important is the spectral reflectance of Earth's surface, importantly including biological influences. These principal elements of Earth's climate are treated in Refs. 6–13. Set the system in motion and it evolves according to the several physical laws. That said, however, the climate system is highly nonlinear, with resultant strong sensitivity of its evolution to even slight changes in initial conditions.[39] This sensitivity limits the accuracy of any calculation of future state of the climate, necessitating solutions that are not expected to be accurate for a given date in the future but are thought to be accurate in the mean and other statistical properties. This statistical independence of initial conditions, which rests on a conjecture by Lorenz,[40] is widely assumed by climate scientists but is not proved and is the subject of much attention by mathematicians.[41]

The principal constituents of the atmosphere are nitrogen, N_2, (78%), oxygen, O_2, (21%), and argon (1%) as mixing ratio (mole fraction) in dry air. The mixing ratio of water vapor is highly variable, from as great as 3% down to $10^{-4}\%$, as governed by temperature and relative humidity. Other gases, present in trace amounts, play no appreciable role in the physical properties of the atmosphere such as density and viscosity, but play key roles in climate because of their influences on absorption and emission of radiation.
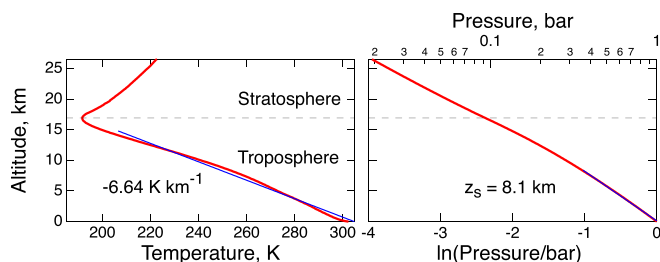
Fig. 5. Vertical structure of atmospheric temperature and pressure in the lower atmosphere (by convention the independent variable altitude is shown on the *y*-axis). Temperature profile is annual mean at the Department of Energy Atmospheric Radiation Measurement site at Manus Island, Papua New Guinea (Ref. 42). Temperature decreases nearly linearly with altitude in the troposphere to a minimum (dashed line) above which it increases; logarithm of pressure *p* decreases approximately linearly with altitude. Straight lines (blue color online) indicate slopes of linear fits over indicated ranges; scale height is evaluated from slope as $z_s = -(d\ln p/dz)^{-1}$.

For the purpose of this Resource Letter, certain properties of Earth's present atmosphere and climate system are taken as givens, although some can be derived.[9,10,13] Key among physical properties of the atmosphere is temperature structure in the lowest two major compartments of the atmosphere, which are most pertinent to climate change: the troposphere and the stratosphere (Fig. 5). The troposphere is more or less well mixed by convection. The rate of decrease of temperature with altitude in the troposphere (denoted lapse rate), about 6.5 K·km⁻¹, is governed by the condition of neutral stability against vertical motion taking into account the latent heat of water condensation in addition to the heat capacity of air. The increase in temperature with increasing altitude in the stratosphere is due mostly to absorption of solar radiation, mainly by ozone; as warmer air is lighter than cooler air, the stratosphere is stable against vertical motion. The altitude of the tropopause, the boundary between the troposphere and the stratosphere, is about 16 km in the tropics, Fig. 5, decreasing poleward to about 11 km in midlatitudes and 9 km at the poles.

Atmospheric pressure decreases from its value at the surface $p(0)$ nearly exponentially with altitude, z, as

$$p(z) = p(0)e^{-(Mg/R)\int_0^z (1/T(z'))dz'} \quad (6)$$

where M is the average molecular weight (molar mass) of air, 0.029 kg·mol⁻¹, g is acceleration of gravity, and R is the gas constant; here the ideal gas law has been employed to calculate mass concentration. The height of the atmosphere if it were compressed to sea-level pressure, 1 bar (1×10^5 Pa), a measure of the scale height of the atmosphere, is about 8 km. If the atmosphere were isothermal, the scale height would be equal to $RT/Mg = 8.4$ km for $T = 287$ K. The troposphere and the stratosphere are more or less decoupled, as the increase in temperature with altitude in the stratosphere suppresses convective transport between the two layers. The circulation between the troposphere and the stratosphere is driven mainly by convection in the tropics and return of stratospheric air to the troposphere at high latitudes.

In addition to the climate system as a whole being in near steady state, each of its compartments is likewise in near steady state. Hence it is required that in each compartment uptake of energy via one process be balanced by loss via another process; estimates of the several energy fluxes shown

in Fig. 2 are constrained by this requirement. Trace atmospheric constituents play essential roles by absorption and emission of radiation in maintaining the atmospheric and surface temperature. The energy budget of the stratosphere consists of absorption of solar radiation in the ultraviolet, mainly by ozone (O_3), balanced by emission in the thermal infrared, mainly by carbon dioxide (CO_2). Although the atmosphere is highly transmissive of solar radiation, near infrared absorption contributes to the energy budget of the troposphere. Solar radiation incident on and absorbed by the surface heats the surface; the surface loses this energy mainly by radiating in the thermal infrared, as a nearly black body. Absorption of this thermal infrared energy heats the adjacent atmosphere; this heating induces convection that tends to mix the troposphere vertically. In these two ways, radiative heating is responsible to first order for the vertical structure of atmospheric temperature. Energy is lost from the atmosphere mainly by thermal infrared radiation, to the surface or to space.

Another key compartment of the climate system is the world ocean. There is much more mass in the water of the ocean than in the air in the atmosphere; the mass of an atmospheric column is equal to the mass of the top 10 m of a column of seawater. Further, the specific heat of water is several times greater than that of air. Consequently, the heat capacity of an atmospheric column is equal to the heat capacity of a column of seawater that is just 2.5-m deep. The ocean is more or less well mixed to a depth of about 100 m, roughly the depth of penetration of the seasonal variation of air temperature. The atmosphere rapidly exchanges heat with the upper ocean, on a time scale less than a year; the exchange of heat between the upper ocean and the deep ocean is much slower than mixing in the upper ocean. Therefore, on the time scale of a decade or so, the effective heat capacity of the climate system is that of the top 100–200 m of the ocean, whereas on longer time scales the rate of climate change is governed by the rate of transfer of heat to the deep ocean. The world ocean thus serves as the flywheel of the climate system, tending to damp rapid fluctuations in the temperature of the atmosphere. In contrast, the change in heat content of land is much less important because of the lower specific heat and lower rate of heat transfer in solid Earth.

The physical, chemical, and biological processes that control Earth's climate must be represented in climate models, numerical representations of these processes. Such models are developed at a hierarchy of complexity, from energy-balance models that give only a coarse representation of these processes to large-scale computer models that represent spatial and temporal (diurnal, seasonal) variation in many climate variables (wind, cloudiness, and precipitation) that are important in climate and climate change. Such models have resolution typically of 100 km or so in the horizontal and twenty layers each in the atmosphere and the ocean, thereby providing much detail. These models do a remarkably good job at reproducing many features of Earth's present climate. Development and application of these large-scale computer models constitutes a major endeavor in current climate research with periodic intercomparisons of the models as to their performance in representing present and prior climate and changes that would result from prospective future changes in emissions.[43] A strength of the energy-balance models is their ability to provide a transparent representation of the large-scale processes and the underlying physical principles that govern the climate that is often

difficult to get from large-scale models.[44] So called Earth system models of intermediate complexity reduce the number of processes and/or the detail of description, for example, to study responses to multiple alternative emissions strategies, with greater treatment of spatial and temporal variability than can be achieved with energy balance models and closer fidelity to large-scale climate models.[45]

39. "Deterministic nonperiodic flow," E. N. Lorenz, J. Atmos. Sci. **20**, 130–141 (1963). http://journals.ametsoc. org/doi/abs/10.1175/1520-0469(1963)020%3C0130:DNF%3E2.0. CO;2. Historically important paper on chaos in atmospheric science. (A)
40. "Predictability; Does the flap of a butterfly's wings in Brazil set off a tornado in Texas?," E. N. Lorenz, American Association for Advancement of Science, Talk presented Dec. 29, Section on Environmental Sciences, New Approaches to Global Weather: GARP. Boston, MA (1972). http://web.mit.edu/lorenzcenter/about/LorenzPubs/Butterfly_1972.pdf. Historically important and accessible. (E)
41. "The Lorenz Attractor, A Paradigm for Chaos," É. Ghys, in Chaos (pp. 1–54), Editors: B. Duplantier, S. Nonnenmacher, and V. Rivasseau (Springer Basel, 2013). https://link.springer.com/chapter/10.1007/978-3-0348-0697-8_1. Mathematical. (A)
42. "Response of tropical precipitation to global warming," D. M. Romps, J. Atmos. Sci. **68**, 123–138 (2011). http://journals.ametsoc.org/doi/abs/10.1175/2010JAS3542.1. Cited as providing vertical temperature profiles. (I)
43. "An overview of CMIP5 and the experiment design," K. E. Taylor, R. J. Stouffer, and G. A. Meehl, Bull. Am. Meteorol. Soc. **93**(4), 485–498 (2012). http://journals. ametsoc.org/doi/abs/10.1175/BAMS-D-11–00094.1. (I)
44. "Energy balance climate models," G. R. North, R. F. Cahalan, and J. A. Coakley, Revs. Geophys. **19**, 91–121 (1981). http://onlinelibrary.wiley.com/doi/10.1029/RG019i001p00091/full. Insightful treatment of heat transfer in the climate system. (I)–(A)
45. "Earth system models of intermediate complexity: Closing the gap in the spectrum of climate system models," M. Claussen *et al.*, Clim. Dyn. **18**(7), 579–586 (2002). https://link.springer.com/article/10.1007/s00382-001-0200-1. (I)

IV. THE GREENHOUSE EFFECT

A. Processes comprising the greenhouse effect

As noted in Sec. I, the greenhouse effect is central to the climate system of Earth (and also of Mars and Venus[46]). Radiative effects of GHGs play a central role in climate change because of the increase in the greenhouse effect resulting from increases in amounts of atmospheric GHGs over the Anthropocene. It is thus essential to understand the greenhouse effect and its dependence on properties of the atmosphere. As the greenhouse effect is a consequence of absorption and emission of longwave radiation by an atmospheric constituent it depends on the local absorptivity (equal, by Kirchhoff's law of thermal radiation, to emissivity) and the local atmospheric temperature. Absorptivity is governed by the amount and wavelength dependent absorption coefficients of the greenhouse gases and other greenhouse agents (clouds and, to much lesser extent, aerosols).

The processes that comprise the greenhouse effect are illustrated schematically in Fig. 6. Panel (a) shows the situation in the absence of an infrared-active gas (or other absorbing substance) in the atmosphere. Thermal radiation is emitted at the surface. As there is no absorption or emission in the atmosphere, the upwelling longwave irradiance at the TOA is equal to the irradiance emitted at the surface. Panel (b) introduces a hypothetical downwelling emission source at the TOA at the same temperature as the surface, but still without an infrared-active gas in the atmosphere. Because the emitted flux from this hypothetical source is equal and opposite to the emitted flux at the surface, there is no net irradiance at the TOA. As all fluxes are matched by equal fluxes in the opposite direction, the system is in a state of dynamic equilibrium. The hypothetical source at the TOA is maintained in Panel *c*, to which a slightly absorbing infrared-active trace gas has been added to the atmosphere. The entire system is isothermal at the surface temperature and the system is still one of equilibrium. (The condition of the gas, or other substance, being optically thin is not essential here, but
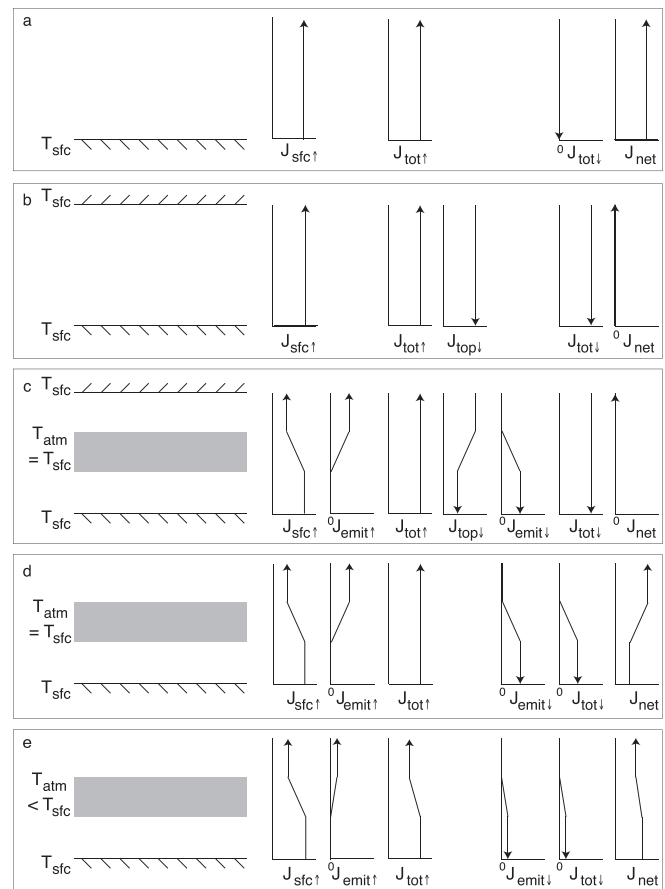

Fig. 6. Absorption and emission of thermal infrared radiation for situations denoted by sketches at left: (a) No absorbing gas. (b) No absorbing gas, but with hypothetical downward-emitting surface at top of atmosphere at temperature of surface. (c) As (b), but with absorbing gas, denoted by gray band, present at temperature T_{sfc}. (d) As (c), but no emitting surface at top of atmosphere. (e) As (d), but gas is at temperature T_{atm} less than T_{sfc}. For each situation longwave radiative fluxes are shown schematically as function of height in atmosphere, vertical axis. The horizontal axis indicates the magnitude of the several components of flux J: upwelling (↑) or downwelling (↓), emanating from surface (sfc), emitted by gas (emit), or emanating from hypothetical downward-emitting surface at the top of the atmosphere (top); total upwelling or downwelling (tot); and net, upwelling minus downwelling; T denotes temperature at surface (sfc) or in atmosphere (atm).

it avoids the complication of the radiation emitted by the gas being subsequently absorbed during its traverse through the atmosphere.) Radiation emitted at the surface is absorbed as it traverses the gas, diminishing this flux. At the same time, the gas is emitting a flux in the upward direction that is equal to the absorbed flux. This emitted flux cancels the decrease in upward flux due to absorption, so that the total upward irradiance is a constant with height. Likewise, the component of the downward irradiance from the hypothetical source is decreased by absorption, in exactly the same amount as the absorption of the upward flux. As with the upward flux, this absorption is compensated by a downward emitted flux from the gas, so that the total downward irradiance is constant with height as well. The net irradiance is zero at all heights, as it must be for an equilibrium system (equal and opposite fluxes on all paths).

In Panel (d), the infrared-active gas is still at the temperature of the surface but the hypothetical source at the TOA is no longer present. The upwelling irradiance is the same as in Panel *c*, (and indeed as in all the preceding panels) as this upwelling irradiance is unaffected by the downwelling irradiance. The absorption is compensated by emission so that the total upwelling flux is constant with height. There is no downwelling irradiance from the top but there is still a downward component of the irradiance emitted by the gas, the same as in Panel (c), as this emitted irradiance does not depend on the irradiance from the top but is a consequence only of the temperature and the emission properties of the gas. The downward emitted irradiance is equal to the upward emitted irradiance, which in turn is equal to the decrease in the upwelling flux from the surface due to absorption. Consequently, there is a component of downwelling irradiance at the surface that would not be present in the absence of the gas (compare Panel (a)), and a resultant decrease in the net irradiance at the surface. In this isothermal situation there is a surface greenhouse effect but no greenhouse effect at the TOA. It might be observed that more energy is exiting the system, than is entering it. Where does this energy come from? It is drawn from the heat reservoir of the atmosphere thereby resulting in a cooling effect locally that is due to the net radiation by the trace infrared-active gas. The situation is commonly characterized as one of local thermodynamic equilibrium. It is *near* thermodynamic equilibrium, but it is not a true equilibrium, as fluxes are not equal and opposite on all paths; it is a steady state. As the GHGs are extracting energy from the surrounding heat bath they must be (very slightly) cooler than the non-radiatively active gases that comprise the bulk of the local atmosphere.

Finally, Panel (e) depicts the non-isothermal situation, in which the infrared-active gas, higher in the atmosphere than the surface, is at a temperature lower than that of the surface. The absorption of surface irradiance is exactly the same as in Panels (c) and (d), as the absorption does not depend on the temperature of the gas. However, as the temperature is lower than in those cases, the thermal emission by the gas is reduced and no longer fully compensates the absorption of upwelling surface irradiance. Consequently, there is a reduction in the upwelling irradiance at the TOA from what it would be if the gas were at the same temperature as the surface (d) or if the gas were entirely absent (a). This is the top-of-atmosphere greenhouse effect. Emission from the gas in the downward direction is likewise less than in the isothermal situation; the net irradiance from the surface is still diminished from what it would be in the absence of the gas

(a) but not so much as in the isothermal situation (d). The net effect of the absorption and emission of radiation is local radiative cooling of the atmosphere, or at a temperature sufficiently low that emission is less than the power absorbed from below, radiative heating of the atmosphere. In the real world, maintaining steady state in the system requires that the energy that is radiated from the atmosphere in the longwave be restored. This is achieved in the troposphere mainly by convection of latent and sensible heat and in the stratosphere mainly by absorption of shortwave energy.

In summary the greenhouse effect induced by infrared-active gases consists of three elements.

- Downwelling longwave irradiance from the atmosphere to the surface, heating the surface and elevating the surface temperature;
- A decrease in upwelling irradiance at the TOA; and
- Radiative cooling and heating of the atmosphere, affecting the vertical temperature structure of the atmosphere.

The greenhouse effect at the TOA consists of two terms: absorption, which results in a decrease in outgoing longwave radiation; and emission, which results in an increase that is subtractive from the first term; As examined in an example developed in Resource Letter GECC-2, the two terms are comparable in magnitude, with the distribution between the two terms depending on the energy of the transition, but with the absorption term dominant.

B. Greenhouse gases in Earth's atmosphere

The principal greenhouse gases in Earth's atmosphere, in addition to water vapor, are carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), and chlorofluorocarbons (CCl_2F_2, CCl_3F), all of which are long-lived in the atmosphere (decade to century). Such lifetimes are much greater than the time required for a gas to become well mixed within the atmosphere ($\sim$3 weeks in a given latitude band, $\sim$3 months hemispherically, 1–2 years globally). Hence, these gases are quite uniformly distributed in the atmosphere, with variations typically 10% or less, Fig. 7. CO_2 exhibits seasonal variation due to seasonal uptake and release by terrestrial vegetation; this seasonal variation is greater in the Northern Hemisphere than the Southern Hemisphere because of greater land mass. Locally, near emission sources or in poorly ventilated vegetated canopies, CO_2 can exhibit considerably greater variation. Importantly, as examined in Resource Letter GECC-2, the atmospheric amounts of these long-lived GHGs, which are usefully considered controlled by processes external to the climate system, have increased substantially over the Anthropocene; such increases, evident even in the few years encompassed in Fig. 7(b), are the basis for concern over the human-induced increase in the greenhouse effect.

It is important to distinguish water vapor from the long-lived GHGs. Water in the atmosphere is present primarily as vapor (even in clouds). An effective upper limit to the partial pressure of water vapor partial pressure is set by the local temperature and the Clausius-Clapeyron equation. Over the global range of surface temperatures, −40 to +35 °C, the equilibrium partial pressure of water vapor varies by a factor of 400, and when the range is extended to tropopause temperature, 210 K, a further factor of 20.

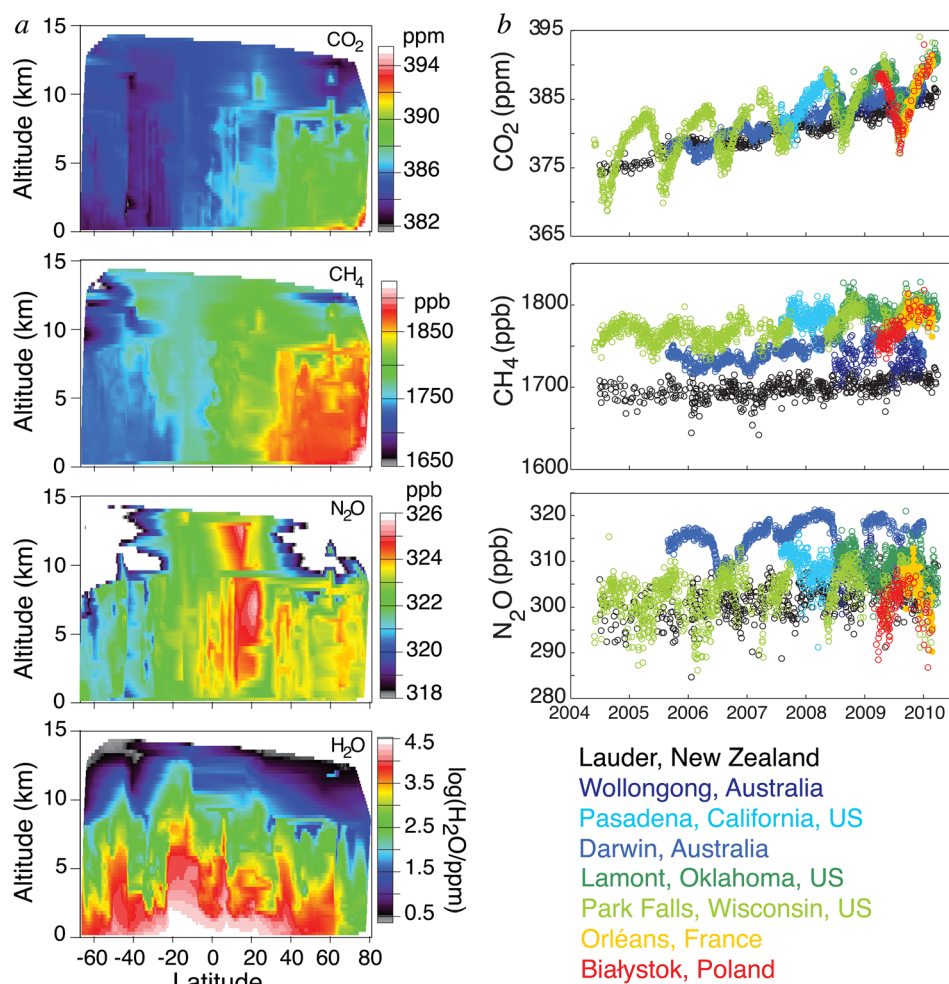The sources of atmospheric water vapor, evaporation of liquid water (and sublimation of ice) and transpiration

Fig. 7. Space and time variation in atmospheric greenhouse gases in the troposphere and lower stratosphere. (a) Mixing ratios of CO_2, CH_4, and N_2O, and water vapor (log scale) along the date line, January 2009. Modified from Ref. 48. (b) Time series of daily-median column-averaged mixing ratios of CO_2, CH_4, and N_2O at several sites. Modified from Ref. 49. Mixing ratio, denotes mole fraction relative to dry air, generally expressed as ppm (parts per million) or ppb (parts per billion).

through the leaves of vegetation, are highly variable in space and time. The rate of evaporation of water from liquid surfaces and moist soil depends strongly on temperature, relative humidity, wind speed, and local turbulence. The rate of transpiration is affected also by sunlight intensity and by water stress, which induces stomatal closure in plants.[47] The principal process whereby water is removed from the atmosphere is precipitation, which is intermittent, with condensation (dew, frost) a distant but non-negligible second. The release of latent heat as water vapor condenses warms the air, enhancing buoyant convection, and thus plays a large role in the dynamics and vertical transport of the atmosphere. Conversely, as air is warmed by compression during downward motion, evaporation of water cools the air, enhancing the downward motion or suppressing convection. Water vapor is short-lived in the atmosphere compared to other major GHGs, with a mean residence time, estimated as the typical amount of water in the atmosphere (equivalent to 2.5 cm of liquid water) divided by the precipitation rate (1 m·yr^{-1}) of 0.025 yr, or 9 days. Such a short residence time, together with the spatial variability of sources and temporal intermittency of removal processes, contributes further to the large spatial, Fig. 7(a), and temporal variability of water vapor in the troposphere. Consequently, it is useful to consider water part of the climate system.

The radiative effects of GHGs, including water vapor, can be calculated using models of varying complexity, depending on application.[46] The most detailed models explicitly represent infrared transitions on a line-by-line basis that includes representation of altitude dependence of collisionally induced broadening and representation of the water vapor continuum. A database of absorption line strengths of atmospheric gases is maintained for this purpose.[50,51] Also necessary inputs to the radiation-transfer calculations are the mixing ratios of the several GHGs and the vertical structure of temperature and water vapor. These radiation-transfer models have been extensively evaluated under conditions for which the atmospheric structure of temperature and water vapor is accurately and independently measured. Departures of modeled and observed downwelling clear-sky fluxes, over a wide range of atmospheric states, are less than 1.5 W·m^{-2}.[52] With confidence in these models, they can serve as the basis for formulating rapid radiation-transfer models that can be employed for calculating radiative effects of GHGs in climate models.[53,54] Such radiation transfer models can also be used to examine questions such as the relative contributions of various atmospheric constituents to the greenhouse effect. About 50% of the total greenhouse effect in Earth's current atmosphere is attributed to water vapor, 25% to clouds, and 20% to CO_2, with the remainder

due to methane, nitrous oxide, ozone, and chlorofluorocarbons, with about 1% contribution from aerosols.[55]

The strong coupling of the abundance of GHGs to Earth's climate is demonstrated in correlations between mixing ratios of CO_2 and CH_4 with other indicia of climate change, the ratios of trace isotopes D and ^{18}O to the principal isotopes H and ^{16}O in glacial ice cores. The most informative ice cores have been obtained in Antarctica, where there is an undisturbed record extending back 800,000 years, **SN3**. These correlations are perhaps the strongest body of evidence of the coupling between GHG abundance and climate state.

46. "Infrared radiation and planetary temperature," R. T. Pierrehumbert, Phys. Today **64**(1), 33–38 (2011). Lucid blend of observation and theory. (E)

47. "Measuring H_2O and CO_2 fluxes at field scales with scintillometry: Part II—Validation and application of 1-min flux estimates," B. Van Kesteren, O. K. Hartogensis, D. van Dinther, A. F. Moene, and H. A. R. De Bruin, Agric. For. Meteorol, **178**, 88–105 (2013). https://www.sciencedirect.com/science/article/pii/S0168192313000154. Novel field measurements. (I)

48. "HIAPER Pole-to-Pole Observations (HIPPO): Fine-grained, global-scale measurements of climatically important atmospheric gases and aerosols," S. C. Wofsy and The HIPPO science team and cooperating modellers and satellite teams, Philos. Trans. R. Soc. A: Math. Phys. Eng. Sci. **369**, 2073–2086 (2011). http://rsta.royalsocietypublishing.org/content/369/1943/2073.short. Simultaneous measurements of multiple trace constituents. (E)

49. "The total carbon column observing network," D. Wunch *et al.*, Philos. Trans. R. Soc. A: Math. Phys. Eng. Sci. **369**, 2087–2112 (2011). http://rsta.royalsocietypublishing.org/content/369/1943/2087.short. Field measurements. (E)

50. "The HITRAN 2012 Molecular Spectroscopic Database," L. S. Rothman *et al.*, J. Quant. Spectrosc. **130**, 4–50 (2013). (I).

51. "HITRAN online: An online interface and the flexible representation of spectroscopic data in the HITRAN database," C. Hill *et al.*, J. Quant. Spectrosc. Radiat. Transfer **177**, 4–14 (2016). This and the preceding reference present spectroscopic data for use in radiation-transfer modeling. (I).

52. "The QME AERI LBLRTM: A closure experiment for downwelling high spectral resolution infrared radiance." D. D. Turner *et al.*, J. Atmos. Sci. **61**, 2657–2675 (2004). http://journals.ametsoc.org/doi/abs/10.1175/jas3300.1. Comparison of measured and modeled spectral downwelling irradiance in the infrared. (I)

53. "Atmospheric radiative transfer modeling: A summary of the AER codes," S. A. Clough *et al.*, J. Quant. Spectrosc. Radiat. Transfer **91**, 233–244 (2005). https://www.sciencedirect.com/science/article/pii/S0022407304002158. (A)

54. "Radiative forcing by long-lived greenhouse gases: Calculations with the AER radiative transfer models," M. J. Iacono *et al.*, J. Geophys. Res. **113**, D13103 (2008). This and the two preceding references are examples of state-of-the-art modeling of atmospheric radiation transfer. (A)

55. "The role of long-lived greenhouse gases as principal LW control knob that governs the global surface temperature for past and future climate change," A. A. Lacis, J. E. Hansen, G. L. Russell, V. Oinas, and J. Jonas, Tellus B, 65 (2013). Accessible treatment of greenhouse effect of long-lived gases and water vapor. (I)

V. SUMMARY

This Resource Letter has examined the energy budget of Earth's climate system, calling attention to pertinent primary and secondary literature. The principal terms in this budget are absorption of shortwave (solar) radiation and emission of long-wave (thermal infrared) radiation. Global mean shortwave absorption, quantified as global average at the top of the atmosphere, is the difference between incident solar irradiance, $340 \, \mathrm{W \cdot m^{-2}}$ and reflected shortwave irradiance, $100 \, \mathrm{W \cdot m^{-2}}$. This absorbed power is balanced by upwelling longwave irradiance, $240 \, \mathrm{W \cdot m^{-2}}$, likewise global average. This emitted radiative flux is much less than the longwave irradiance emitted at the surface (about $385 \, \mathrm{W \cdot m^{-2}}$, global average). The difference between these fluxes is a consequence of the so-called greenhouse effect, absorption of longwave radiation by trace species in the atmosphere, water vapor, CO_2 and other polyatomic molecules, and clouds. This difference in fluxes results in a return, longwave flux to the surface that is responsible for the global mean temperature at the surface, average roughly 287 K, being much greater than the radiative temperature at the top of the atmosphere, 255 K. The greenhouse effect is thus a central feature of Earth's climate system.

It must be stressed that the several fluxes given above are averages of quantities that exhibit large spatial and temporal variation, a consequence of diurnal and seasonal variation in solar irradiance and also of large variation in atmospheric and surface composition, mainly variation in the amount of water, in the form of vapor, clouds, and snow and ice, that greatly affects absorption of longwave radiation and scattering and reflection of shortwave radiation. This variation in the several fluxes presents large challenges in quantifying Earth's radiation budget. In view of measurement uncertainty, the assertion of balance between absorbed shortwave radiation and emitted longwave radiation rests not so much on measurement as on the consequence of an imbalance, which would be manifested by rapid increase or decrease in planetary temperature, The examination of Earth's radiation budget presented here sets the scene for the examination in Resource Letter GECC-2 of the perturbation in this budget due to increases in the amounts of the several greenhouse gases over the Anthropocene and the consequences of these increases for Earth's climate.

ACKNOWLEDGMENTS